On Aerosol and Molecular Extinction of Solar Radiation in Transparency "Window" ν = 750 - 1250 cm^{-1} from Atmospheric Measurements

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Introduction

In investigations of variations of optical depth of the vertical column of atmosphere $\tau = -\ln P$, where P = transmittance of the atmosphere, in intervals $\Delta\nu_i$ of transparency "microwindows" i in range of wavenumbers ν = 750 - 1250 cm^{-1}, one of the most difficult problems is to separate the contributions of an aerosol and water vapor in τ (spectral position of "microwindows" i (see Figure 1 under numbers i = 1 - 22). Optical depth τ (m, w), measured at air mass m and amount of the precipitable water vapor H_2O in the vertical column of atmosphere w (cm), is possible to express by the sum (Bignell, Saiedy, and Shepard 1963):

$$\tau(m) = \tau_a(m) + \tau_{mol}(m) + \tau^k{}_{H_2O}(m) + \tau^c{}_{H_2O}(m) \quad (1)$$

where:

τ_A (m) = aerosol optical depth
P_A (m) = exp [$-\tau_A$ (m)] - transmittance of an aerosol
τ_{mol} = optical depth of gases N_2O, CO_2, CH_4, O_3 (the contribution in $\Delta\nu_i$ of other gases can be neglected)
P_{mol} (m) = exp [$-\tau_{mol}$ (m)]
$\tau^k_{H_2O}$ (m) = $k^k_{H_2O} \times w \times m$
$\tau^c_{H_2O}$ (m) = $k^c_{H_2O} \times w \times m$ - continuum and selective components of a water vapor optical depth
$k^k_{H_2O}$ and $k^c_{H_2O}$ = corresponding coefficients of the absorption H_2O (cm^{-1})
$P^k_{H_2O}$ (m) = exp [$-\tau^k_{H_2O}$ (m)]
$P^c_{H_2O}$ (m) = exp [$-\tau^c_{H_2O}$ (m)].

At present, the laboratory investigations have not revealed the nature of variations of dependence $\tau^k_{H_2O}$ upon the water vapor partial pressure e (mb) and air temperature t (°C)

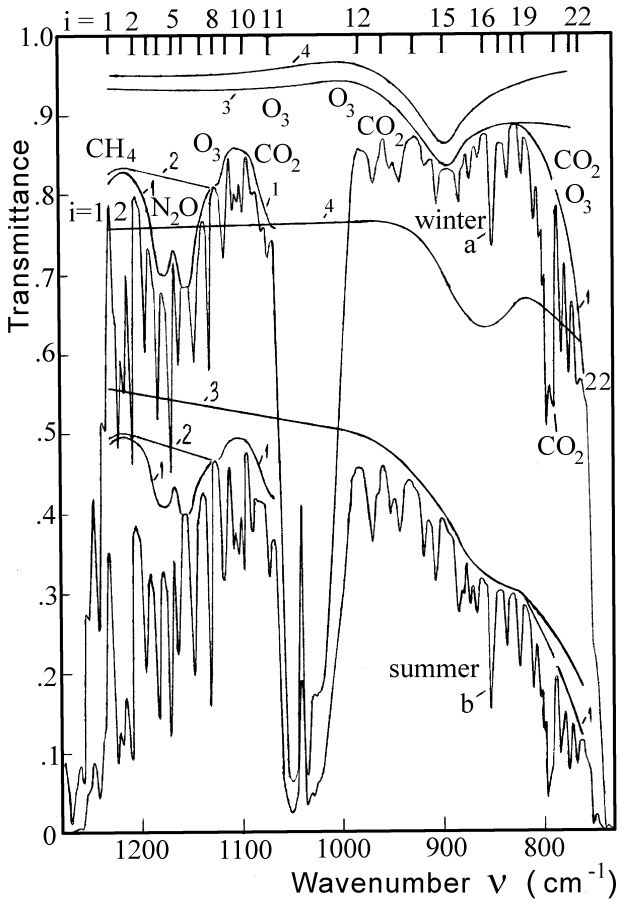

Figure 1. Examples of transmittance P_ν at m ≈ 3: a - winter, b - summer.

(Kulp and Shinn 1994). This is the main difficulty of an estimation of τ_A, which is necessary, particularly, for calculations of the climatic effects of aerosol and clouds. Another difficulty is connected with the application of the Bouguer's law for determination $\tau^k_{H_2O}$, $k^k_{H_2O}$, and their spectral dependencies (Paramonova, Brounstein, and Frolov

1978). The values τ_i in "microwindows" i are estimated, usually, by Bouguer's law:

$$\ln I_i = \ln I_i^o - \tau_i \times m_i \qquad (2)$$

where I_i and I_i^o = measured and top-of-the-atmosphere values of relative intensity I_ν of solar radiation in $\Delta\nu_i$.

Values I were measured for estimations of $\tau_{i,A}$ at optimum spectral resolution $\delta\nu$ = 1 - 5 $cm^{-1} \approx \Delta\nu_i$. Using (2) in $\Delta\nu_i$, where $\tau_{mol} \neq 0$ and $\tau^c_{H_2O} \neq 0$, by calculations of Paramonova et al. (1978) leads to underestimation of I_i^o and τ_i, as a result of τ_i (m) $\neq \tau_i \times$ m. It was not experimentally checked, since usually I_ν were measured at w $\geq$ 0.5 cm, which didn't permit the separation of molecular bands, while at small values of w ($\leq$ 0.3 cm) they are discovered quite clearly (compare spectra P_ν on Figure 1).

The goal of the present study is to obtain distinctions $P_{\nu,A}$ in winter and summer conditions, highly different by w.

Data and Method

The spectra of $P_\nu = I_\nu / I_\nu^o$, obtained in different seasons of 1971 - 1973 at Zvenigorod Scientific Station of IAPh RAS, were used for the estimations of τ and its components. I_ν^o was calculated by interpolation of values I_i^o obtained from (2) with subsequent adjustment by the recommendations of Paramonova et al. (1978). Spectra of I_ν were registered with a two-beam prism spectrophotometer UR - 20 with $\delta\nu$ = 4.5 cm^{-1} and a scanning rate of 40 cm^{-1} / min. Over 250 realizations I_ν were recorded at variations w within the range from $\approx$ 0.2 cm to $\approx$ 2.6 cm and at near-surface temperature t^o from – 25 °C up to + 30 °C. Over 100 of them corresponded to w $\leq$ 0.3 cm and $t^o < 0$. Calibration dependencies of P_{min} (m × w)—transmittance in maximum of 2724 cm^{-1} HDO absorption band related to transparency "microwindows" in the region of its wings—were used for estimation of w, which reduced the error of the calculation of w up to ± 5%. The method of the estimation of components of τ in typical winter and summer conditions is shown in Figure 1 for

$$P_\nu = P^c_{\nu,H_2O} \times P_{\nu,mol} \times P^k_{\nu,H_2O} \times P_{\nu,A}$$

As an example: a - winter, w = 0.23 cm, $t^o \approx$ – 10 °C, m = 3.27 (i = 1) - 3.46 (i = 22); b - summer, w = 2.1 cm, $t^o \approx$ + 25 °C, m = 3.11 - 3.36.

The lines 1 - 4 above P_ν - (a) were plotted according to estimates of the contribution of gas components (above P_ν - (b) they are plotted similarly). The determination of factors P_ν were obtained by the following way. The line 1 is P / $P^c_{H_2O}$. Values k^c_{i,H_2O} in i = 1 - 5, 7, 9 - 11 were estimated by dependence of $c_{i,6} = P_i / P_6$ upon m in P_ν - (a) and P_ν - (b), $k^c_{i,H_2O} \approx 0$ was assumed for other i. Since $k^c_{6,H_2O} \approx 0$,

$$k^c_{i,H_2O} = \Delta k^k_{i,H_2O} + [1/(\Delta w \times m)] \times \ln[c_{i,6}(w_a)/c_{i,6}(w_b)] \qquad (3)$$

where $\Delta k^k_{i,H_2O} = k^k_{i,H_2O} - k^k_{6,H_2O}$ was calculated from spectral dependence of k^k_{i,H_2O} (see below), Δw = 2.1 cm – 0.23 cm $\approx$ 1.9 cm. A line 2 - $P^o_{N_2O}$ (overline of N_2O bands) was plotted by P in i = 8 and by values P / ($P^c_{H_2O} \times P^o_{N_2O}$) in i = 1, 2, where $\tau_{N_2O} \approx 0.005$ atm^{-1} by published data. A line 3 - $P^k = P_A \times P^k_{H_2O}$ = P / ($P^c_{H_2O} \times P_{mol}$) was drawn by interpolation of values P^k in i = 2, 8, 9, 12, 15, 16, 18 and 22 (in i = 15, 16, 18 $P^k \approx P$). $P^k_8 = P_8 / P_{8,O_3}$, $P^k_9 = P_9 / (P_{9,O_3} \times P^c_{9,H_2O})$. Values of P_{O_3} for i = 8 have been estimated at O_3 content of 0.3 atm.cm according to Bartman et al.(1975). In i = 2 value $P^c_{H_2O} \times P_{mol}$ was determined by dependence of P_2 / P_8 upon m. In i = 12 $P^c_{H_2O} / P_{mol} = P^c_{9,H_2O} \times P_{9,mol}$. In i = 22 $P_{22,mol} \times P^c_{22}$ was calculated by dependence of P_{22} / P_{18} upon m with correction for Δk^k and with assumption $P^c_{18,H_2O} \approx P^c_{22,H_2O} \approx 1$. A line 4 - $P_A = P^k / P^k_{H_2O}$ was drawn by interpolation of values $P_{i,A} = P^k_i / P^k_{i,H_2O}$. Values P^k_{i,H_2O} were evaluated by τ^k_{i,H_2O} and its spectral dependence (Shukurov 1986), where it was shown, that if w measured in (cm), then $\tau^k_{i,H_2O} = \beta \times w^{1.25}$. For w = 1 cm, numerically, $\beta = \tau^k_{i,H_2O}$. Values β for range ν = 765 - 1233 cm^{-1} are presented in Table 1.

Table 1. Values β for range ν = 765 - 1233 cm^{-1}.

ν_i (cm^{-1})	1233	1203	1096	987
i	1	2	10	12
β_i	.038	.040	.045	.053
ν_i (cm^{-1})	901	832	765	
i	15	18	22	
β_i	.068	.093	.14	

The aerosol optical depth $\tau_A = -(1 / m) \times \ln P_A$ was determined with an accuracy of ± 15 %.

Results

The aerosol optical depth in winter in all "microwindows" i (except i = 14, 16) does not exceed $\tau_A = 0.015$. In the summer at large w and at the usual haze conditions, it is more significant, than it was believed. Thus, in i = 12, $\tau_{i\,A}$ reaches 0.08. The features of spectral dependence $P_{v,A}$ (its decrease at $\nu \le 1000$ cm^{-1} and minimum in the range of 900 - 850 cm^{-1}) can be explained by extinction of radiation by water droplets and ice crystals.

Obtained from atmospheric measurement values of k^k_{i,H_2O} (w) agree with laboratory $k^k_{i,H_2O} = \tau^k_{i,H_2O} / w$ of Kulp et al. (1994) within an accuracy of ± 10 %, where τ^k_{v,H_2O} was measured at e ≈ 10.7 mb, t = + 24 °C. For instance, in i = 15 (ν = 901 cm^{-1}), k^k_{15,H_2O} = 0.081 cm^{-1} by our data and 0.085 cm^{-1} by laboratory data. It confirms the suggestion that the water vapor H_2O continuum is determined practically by, so-called, "e-type", water vapor absorption of radiation.

The estimates of $P_{i,mol}$ agree with those calculated by Paramonova et al. (1978); however, for i = 6, our value $\tau_{6,N_2O} = -(1 / m) \times \ln P_{6,N_2O}$ (m), where P_{6,N_2O} (m) = $P_6 / P^o_{6,N_2O}$, more and equal to 0.05 atm^{-1} (compared to calculated data, which was carried out at m = 4). The calculated value of τ_{6,N_2O} is equal to 0.028 atm^{-1} by Bignell et al. (1963) and 0.038 atm^{-1} by Paramonova et al. (1978).

In conclusion, it can be noted that the data on solar radiation extinction at small amounts of water vapor content (in middle latitudes - in winter) are very important for analysis of atmospheric transmittance.

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